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Some controls on flow and salinity in Bering Strait

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[1] During 1993–1994, steric forcing of flow through Bering Strait represented a northward sea level drop of ~0.7 m from the Bering Sea Basin to the adjacent deep Arctic Ocean, of which ~2/3 was due to the salinity difference between the basins. Seasonal variability of steric forcing appears small (<0.05 m), in contrast to large seasonal wind effects. Interannual changes in steric forcing may exceed 20%, however, and warm inflow from the North Atlantic, accumulation of freshwater in the southwest Canada Basin, and temperature and salinity changes in the upper Bering Sea have all contributed to recent changes. The mean salinity balance in Bering Strait is primarily maintained by large runoff to the Bering shelf, dilute coastal inflow from the Gulf of Alaska, and on-shelf movement of saline and nutrient-rich oceanic waters from the Bering Sea Basin. In Bering Strait, therefore, both the throughflow and its salinity are affected by remote events. **Citation:** Aagaard, K., T. Weingartner, S. L. Danielson, R. A. Woodgate, G. C. Johnson, and T. E. Whitledge (2006), Some controls on flow and salinity in Bering Strait, *Geophys. Res. Lett.*, *33*, L19602, doi:10.1029/2006GL026612.

1. Introduction

[2] Northward flow of low-salinity Pacific waters through Bering Strait is a major link in the global water cycle [*Wijffels et al.*, 1992], and changes in Bering Strait flow have been suggested to exercise substantial control over northern hemisphere climate [*Shaffer and Bendtsen*, 1994; *De Boer and Nof*, 2004]. The flow shows pronounced annual cycles in temperature, salinity, and transport: by late winter, water over the northern Bering shelf is near freezing, and the salinity exceeds summer–fall values by 1 (PSS-78) or more, while the mean winter transport is less than half that during summer [*Woodgate et al.*, 2005a]. The Pacific waters subsequently enter the Arctic Ocean, where their salinity is critical to halocline ventilation [*Aagaard et al.*, 1981]. Both seasonal and interannual changes in salinity (which primarily determines the density) will affect the fate of the Pacific waters. Furthermore, except along the NW Alaskan coast during some years, the salinity of the Pacific waters does not appear to be greatly modified during the transit of the Chukchi shelf [*Woodgate et al.*, 2005b; *Weingartner et al.*, 2005a], so that the density of these waters may largely be set prior to their passage through Bering Strait. Our interest here is therefore to explore mechanisms controlling transport and salinity in Bering Strait.

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2. Mass Transport

[3] It has long been argued that mean northward flow through the strait is driven by a sea level dipping down toward the north [*Shtokman*, 1957; *Coachman and Aagaard*, 1966], whether steric in origin [*Stigebrandt*, 1984] or associated with wind-driven circulations [*Gudkovich*, 1962]. *Woodgate et al.* [2005b] suggest that the pressure head may vary seasonally and interannually. We here concentrate on what recent measurements suggest about steric effects, recognizing that the Bering Strait flow is also considerably modified by the wind, which in the mean opposes the northward motion, and which shows strong seasonality [*Aagaard et al.*, 1985; *Roach et al.*, 1995; *Woodgate et al.*, 2005b]. We consider the steric effect only on a regional length scale, i.e., steric height differences between adjacent deep basins in the Bering Sea and the Arctic Ocean, recognizing that sea surface fluctuations, whether of steric or other origin, will also occur on shorter scales over the intervening shelf. The latter fluctuations, however, will not affect the regional mean sea surface slope.

[4] The deepest continuous pressure surface connecting the North Pacific and the Arctic Ocean lies near 800 m (through the Faroe Bank Channel). This isobaric surface (~800 db) grounds in the northern Bering Sea and on the Chukchi slope of the Arctic Ocean. A comparison of deep casts in the northern Bering Sea (stations 3–5 from WOCE section P14N, summer 1993 [*Roden*, 1995]) and in the Arctic Ocean north of the Chukchi Sea (stations 9, 11, and 13 from the Arctic Ocean Section, summer 1994 [*Swift et al.*, 1997]) shows a mean geopotential anomaly difference (here and subsequently calculated for 0/800 db) between the two station groupings of $\sim 7\ \mathrm{m^2\ s^{-2}}$, with the water column being fresher, warmer, and less dense to the south. Two-thirds of this steric effect, corresponding to an ~0.7-m steric sea level drop toward the north, is due to the salinity difference, and the remainder to temperature.

[5] How might the steric height difference vary seasonally? While measurements from comparable areas north of the Chukchi Sea are available only from summer, comparison of October 1986 and April 1987 CTD profiles over the Beaufort Sea slope north of Alaska [*Aagaard et al.*, 1988] shows a steric height ~0.05 m greater in October (taken to represent late summer), primarily due to a warmer and fresher upper layer. In the northern Bering Sea Basin, meanwhile, profiling float data from 2001–2006 [http://floats.pmel.noaa.gov/] show a mean seasonal variability in geopotential anomaly of $\sim 0.7\ \mathrm{m^2\ s^{-2}}$ (Figure 1b), corresponding to a seasonal steric height difference of ~0.07 m (higher in summer). The amplitudes and phase suggest that the seasonal

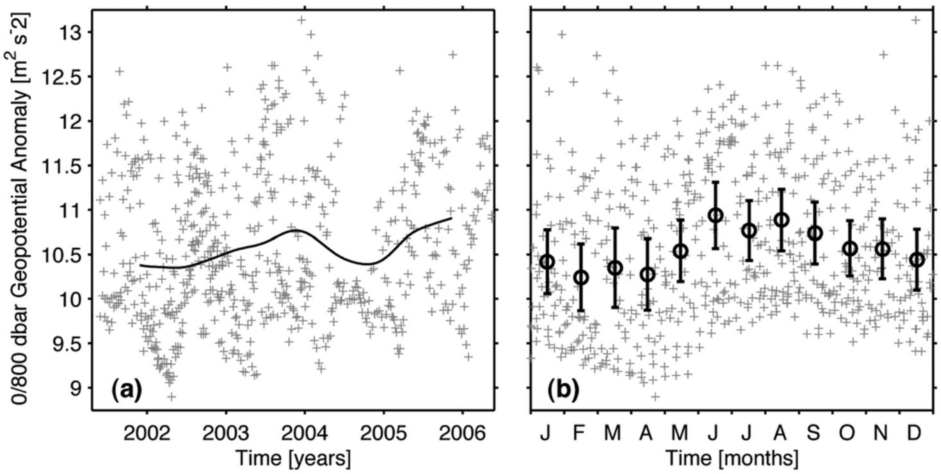

Figure 1. (a) Geopotential anomalies (gray crosses) in the northern Bering Sea, 0/800 db, spring 2001–spring 2005–2006, based on all profiling float observations within the regime of the northwestward flowing Bering Slope Current (compare Figure 1 of *Johnson et al.* [2004]). Anomalies have been corrected for location by an exponential fit of anomaly versus distance from 1000 m isobath. Interannual variations are highlighted with a 1-year half-span loess filter (solid line) fit to anomalies with seasonal cycle removed (not shown). (b) Seasonal distribution of individual anomalies (gray crosses) with monthly averages (circles) and their 95% confidence limits (error bars).

variation in steric height difference between the deep Bering Sea and the adjacent Arctic Ocean is likely small during most years (<0.05 m). Interannual variability may be much larger (Section 4).

[6] There is also a seasonal cycle in the atmospheric pressure difference between the corresponding portions of the Bering Sea and the Arctic Ocean of ∼15 mb (lower pressure to the south in winter, e.g., Figure 7.1 of *Peixoto and Oort* [1992]). This cycle should be compensated through the inverted barometer effect [e.g., *Wunsch and Stammer*, 1997], resulting in little, if any, net contribution to the meridional oceanic pressure gradient from the seasonally varying atmospheric loading.

3. Salinity Forcing

[7] We next consider the mean annual water and salt balances for the Bering shelf (Figure 2). In the north, water and ice are exported through Bering Strait. We take the mean volume transport of liquid sea water through the strait as 27,800 km^3 yr^{-1} (including transport by the Alaskan Coastal Current), and of ice as 100 km^3 yr^{-1} [*Woodgate and Aagaard*, 2005]. Using the mean freshwater transport estimate from their Figure 2, which is relative to a reference salinity of 34.8 and includes the effects of seasonality and stratification, yields a transport salinity of 31.8. For ice we use a salinity of 7 [*Aagaard and Carmack*, 1989].

[8] Runoff onto the Bering shelf is ∼320 km^3 yr^{-1}, of which ∼200 km^3 yr^{-1} comes from the Yukon (gauged at Pilot Station); ∼30 km^3 yr^{-1} from the Anadyr (gauged at Snezjnoje) [http://www.r-arcticnet.sr.unh.edu/]; ∼75 km^3 yr^{-1} from the Kuskokwim, Nushagak, and Kvichak [http://nwis.waterdata.usgs.gov/ak/nwis/nwis]; and perhaps 10–20 km^3 yr^{-1} from other or ungauged streams, including those joining the Anadyr downstream of Snezjnoje.

[9] Additionally, a significant portion of the very large runoff into the Gulf of Alaska (GoA) enters onto the Bering shelf through Unimak Pass. Indeed, *Weingartner et al.* [2005b] argue that this source is a first-order term in the Bering shelf freshwater balance. We take the annual transport through Unimak Pass as 13,000 km^3 yr^{-1} [*Stabeno et al.*, 2002], and if we combine their seasonal distribution of barotropic and baroclinic velocity with the measured seasonal salinity cycle (P. J. Stabeno, personal communication, 2006), we get a transport salinity of 31.7 (rounded down to maintain the ∼0.1 salinity difference between Unimak Pass and Bering Strait indicated by the full calculation).

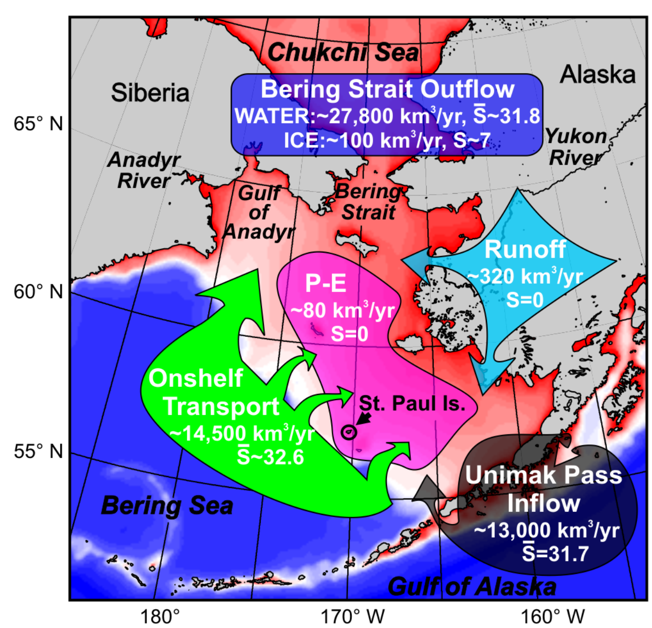

Figure 2. Schematic of the Bering shelf water and salt budgets. Precipitation, evaporation, and salinity are denoted P, E, and S, respectively.

Table 1. Constraints on Water and Salt Budgets[a]

		Bering Strait Transport		
Unimak Pass Transport	Parameter	$T_{BS} - 25\%$	T_{BS}	$T_{BS} + 25\%$
$T_{UP} + 25\%$	$\overline{S}$, H	34.6[b]	32.9[b]	32.5, 70
	S_H			32.8
	$[NO_3]_H$			24
	$[SiO_3]_H$			48
T_{UP}	$\overline{S}$, H	33.3[b]	32.6, 100	32.3, 30
	S_H		32.9	32.4
	$[NO_3]_H$		28	9
	$[SiO_3]_H$		51	10
$T_{UP} - 25\%$	$\overline{S}$, H	32.8[b]	32.4, 50	32.3, 30
	S_H		32.6	32.4
	$[NO_3]_H$		21	9
	$[SiO_3]_H$		26	10

[a]See Figure 2. Salinity and nutrient concentrations (μM kg^{-1}) result from budget calculations with varying transports through Unimak Pass (T_{UP} = 13,000 km^3 yr^{-1}) and Bering Strait (T_{BS} = 27,800 km^3 yr^{-1}). The mean salinity ($\overline{S}$) calculated from the budgets, together with the layer depth (H) required to reach that mean salinity on the southwestern Bering shelf, are shown for each transport combination, as are the salinity (S_H) and nutrient concentrations ($[NO_3]_H$ and $[SiO_3]_H$) observed at that depth.

[b]The mean salinity calculated from the budgets is not observed on the shelf.

[10] We estimate precipitation (P) for the Bering shelf from the St. Paul Island records (see Figure 2 for location), which show a long-term mean of 0.61 m yr^{-1}. This is based on the Cooperative Observer Program station daily summaries archived for 1892–1894 and 1949–2003 by the National Climatic Data Center, together with monthly National Weather Service summaries for 1911–1938 (with a few years missing). Using the mean latent heat flux from the National Centers for Environmental Prediction reanalysis ($\sim$46 W m^{-2}) for the same area during 1948–2005 yields an evaporation (E) estimate of $\sim$0.58 m yr^{-1}. Alternatively, applying *Penman*'s [1948] bulk formula to the St. Paul dew point and air temperature records from the same period gives E$\sim$0.33 m yr^{-1}. Applying the average of these two evaporation estimates uniformly over the entire shelf, together with the precipitation estimate, yields a net P–E of $\sim$80 km^3 yr^{-1}, substantially less than the runoff onto the shelf.

[11] The flux of oceanic water onto the Bering shelf, apart from that entering through Unimak Pass, is unknown, and so we estimate it from the residual in the water and salt balances for the shelf. This yields $\sim$14,500 km^3 yr^{-1} of oceanic waters moving onto the shelf, with mean salinity 32.6, i.e., about 0.8 higher than the mean for the Bering Strait throughflow (Figure 2).

[12] There are large uncertainties in our calculation, e.g., related to the size of the Bering Strait and Unimak Pass transports. The former has estimated errors $\sim$25% [*Woodgate et al.*, 2005a], and for discussion purposes we assume a similar proportional uncertainty in the Unimak Pass transport. We then repeat our calculations of the on-shelf flux and its mean salinity for a range of Bering Strait and Unimak Pass transports within these bounds (Table 1), to afford a comparison with observations over the western Bering shelf, where much of the on-shelf flux is thought to occur [*Coachman*, 1993]. These observed outer shelf values have been calculated from the NODC World Ocean Database 2001 [*Conkright et al.*, 2002] as composite means near 178°W and inshore of the 150-m isobath. In particular, we have calculated the mean salinity from the surface to various depths, together with the salinity and nutrient values at the integration depth (bottom of the layer for which the mean salinity has been calculated). The mean observed water column salinity can then be compared with that calculated from the water and salt budgets, while the salinity and nutrient concentrations at the base of the composite layer (integration depth) that yields a particular mean salinity can be compared with the deep salinity and nutrients observed farther north on the shelf, e.g., in the Gulf of Anadyr. The latter characteristics (S$\sim$33, $NO_3$$\sim$30–35 μM kg^{-1}, $SiO_3$$\sim$55–60 μM kg^{-1}) represent the nutrient-rich source waters for the highly productive western and central Bering and Chukchi shelves [*Walsh et al.*, 1997]. For example, Table 1 shows that an observed mean salinity of 32.6, corresponding to the mean salinity we have calculated from the most likely mass balance (Figure 2), represents a composite water column on the outer shelf 100 m thick, with observed salinity $\sim$32.9, nitrate $\sim$28 μM kg^{-1}, and silicate $\sim$51 μM kg^{-1} at the bottom of that layer. These values are only slightly lower than those observed in the Gulf of Anadyr, consonant with the former representing temporal and spatial means that will underestimate extreme values. In contrast, most of the other transport combinations in Table 1 yield unrealistic water properties. Indeed, only two of those combinations, with mean salinities of 32.4 and 32.5, appear admissible. Table 1 therefore suggests both that our mass budgets (Figure 2) are reasonable, and that a Bering Strait transport much less than our estimate of 27,800 km^3 yr^{-1} is unlikely. (Alternatively, we could have calculated nutrient concentrations at the base of the composite layer from the salinity at that depth, using salinity–nutrient linear regressions. Doing so based on a regression from stations 2–5 of the P14N section (correlation coefficients of 0.5 for nitrate-salinity and 0.6 for silicate-salinity) yields nutrient concentrations very close to the observed composite values for mean salinities of both 32.6 and 32.5, but with poorer agreement for lower mean salinities.)

[13] Calculations (not shown) with mean salinities altered by ±0.2, but with Bering Strait and Unimak Pass transports maintained at 27,800 km^3 yr^{-1} and 13,000 km^3 yr^{-1}, respectively, are also consonant with rather tight bounds on water properties. In particular, they suggest that while the mean transport salinities we adopted for Bering Strait (31.8) and Unimak Pass (31.7) give realistic results for the onshelf flow characteristics to the west, mean salinities in Bering Strait that are much higher than 31.8 are unlikely.

4. Discussion

[14] The seasonal variation in steric sea level difference between the northern deep Bering Sea and the southwestern Canada Basin of the Arctic Ocean is typically <10% of the mean difference. The large annual cycle in transport ($\sim$0.4–1.3 Sv from summer to winter [*Woodgate et al.*, 2005a]) is therefore likely wind forced (compare especially Figure 5 and the associated discussion by *Woodgate et al.* [2005b], although we cannot exclude the possible contribution of an annual cycle in sea level slope that is not steric, e.g., associated with seasonal adjustment of the barotropic circulations in the Bering Sea and the Arctic Ocean.

[15] Are there interannual changes in the steric sea level difference? The profiling float data (Figure 1a) suggest both

that the annual cycle in the Bering Sea may vary from year-to-year and also that there are upper-ocean multi-year trends [cf. also *Wirts and Johnson*, 2005]. Indeed, the trend in the Figure 1 data is equivalent to a steric height increase of ~0.05 m over the record, caused equally by warming and freshening, largely above 300 m. On the other hand, comparing the P14N stations with the float data shows a steric height decrease between 1993 and 2003 of ~0.08 m due to increased salinity. That earlier decadal change is now diminishing rapidly due to the recent freshening and warming.

[16] There is no corresponding time series for the adjacent Arctic Ocean, but a comparison of three stations occupied in late summer 2002 [http://psc.apl.washington.edu/CBL.html] in the same area as those cited for 1994 shows a steric height increase of ~0.08 m (spatial differences among stations during the same year average only 0.02 m). This temporal increase is due both to warming of the Atlantic layer (+0.25°C at 200–600 m) and to marked freshening of the upper ocean (approaching −0.5 above 200 m). The warming is now well documented throughout much of the Arctic Ocean as a propagating anomaly [*Karcher et al.*, 2003], reaching the Chukchi Borderland in the late 1990s [*Shimada et al.*, 2004]. The origin of the freshening in this area is unknown, but may be a combination of runoff and ice melt accumulation [*Macdonald et al.*, 1999].

[17] These results suggest a decrease in steric height difference between the deep northern Bering Sea and the adjacent Arctic Ocean of ~20 % between 1993–1994 and the early part of the present decade, due almost equally to an increase in steric height in the Arctic Ocean because of warming and freshening, and a decrease in the Bering Sea due to salinization. During 2001–2006, however, the steric height in the Bering Sea was again increasing as the upper ocean warmed and freshened. While these various indications are fragmentary, taken together they suggest significant interannual variability in steric height both in the Bering Sea and the adjacent Arctic Ocean, and that the variability in one basin is largely independent of that in the other, resulting in a variable regional sea level slope. In the case examined here, the steric height increase in the Arctic Ocean between 1994 and 2002 resulted from advection of warm water from the North Atlantic together with a regional change in the freshwater distribution, while the steric height decrease in the Bering Sea during about the same period was driven by a change in the salt budget. The variable steric forcing for the Bering Strait was therefore remotely controlled from as far away as the North Atlantic.

[18] Is the current record compatible with this interannual variability? While it is difficult to discern clear transport trends from the Bering Strait current record for 1990–2002, there are suggestions both of a small decrease during 1994–2002 (middle/lower panel of Figure 5 of *Woodgate et al.* [2005a]) and a more rapid increase beginning in 2002 (upper panel of Figure 5 of *Woodgate et al.* [2005a]). Of these suggested changes, only the recent transport increase coincides with a corresponding wind trend, but even then the mean southward wind did not weaken as much as during 1996, when the northward flow was not unusually strong. Wind variability can therefore at best only partially explain interannual changes in the current record, leaving room for variability in the pressure forcing.

[19] Remote forcing also appears critical to the salinity of the waters entering the Arctic Ocean from the Pacific. In particular, the annual mean salinity of the Bering Strait throughflow primarily represents the combined effects of large freshwater sources along the Alaskan coast, both on the Bering shelf and in the GoA, and more saline waters flowing onto the shelf from the Bering Sea Basin. Seasonally, salinity in Bering Strait is modified significantly by the freezing and melting cycle over the Bering shelf, but the net effect on the annual mean salinity in the strait appears small. The mean Bering Strait salinity (and density), and probably also those of the Arctic Ocean halocline [*Weingartner et al.*, 2005a; *Woodgate et al.*, 2005b], are therefore in large part determined by events far to the south, along the Bering shelf edge and in the GoA. In the future, changes in the freshwater fraction transiting the strait would likely first and foremost reflect alterations in these sources.

[20] Finally, the oceanic waters presently moving onto the Bering shelf carry elevated nutrients that fuel high shelf production in both the Bering and the Chukchi seas, and that subsequently can be traced across the Arctic Ocean and into the North Atlantic [*Jones et al.*, 2003]. As with the salinity of the throughflow, the variability of these properties over the northern shelves and within the Arctic Ocean depends in large part on remote forcing that has significant consequences far downstream.

[21] Predicting the evolution of these various fields and their influences, whether on stratification or biological production, will require a considerably better understanding of the regional and large-scale systems than we presently possess.

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